

N94-30625

209932
p. 27

Four Simple Ocean Carbon Models

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The change in atmospheric carbon dioxide concentration resulting from fossil fuel combustion, land management activities, and other human-induced disturbances of the global carbon cycle (Figure 1) is strongly governed by the CO_2 exchange between the atmosphere and ocean. The ocean is believed generally to be the largest sink for atmospheric CO_2 ; however, there is evidence, though quite controversial, that the temperate biosphere may have been a comparable net annual sink during the last decade (Tans et al., 1990).

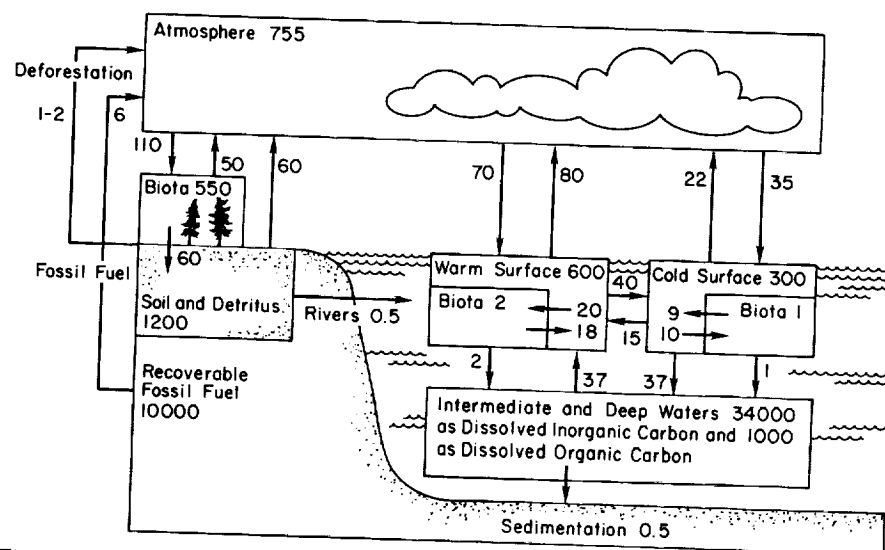


Figure 1. The global carbon cycle. The boxes are in Pg C, and the fluxes (arrows) are in Pg C/yr as CO_2 .

Granting this and other uncertainties about the carbon cycle, future atmospheric concentrations of CO_2 will depend on the rate of industrial CO_2 emissions, the net exchanges between the atmosphere and the terrestrial biosphere, and the rate of uptake by the oceans.

This paper briefly reviews the key processes that determine oceanic CO_2 uptake and sets this description within the context of four simple ocean carbon models. These models capture, in varying degrees, these key processes and establish a clear foundation for more realistic models that incorporate more directly the underlying physics and biology of the ocean rather than relying on simple parametric schemes.

The purpose of this paper is more pedagogical than purely scientific. The problems encountered by current attempts to understand the global carbon cycle not only require our efforts but set a demand for a new generation of scientist, and it is hoped that this paper and the text in which it appears will help in this development. The student reader might also want to study SCOPE Reports 13 (Bolin et al., 1979) and 16 (Bolin, 1981); a review of the carbon cycle is given by Moore (1984); and finally, Moore and Schimel (1992) provide an extremely useful context for these discussions.

Ocean Modeling

The oceanographic community already has a rich history that parallels the atmospheric modelers' advances in seeking the development of ocean general circulation models (GCMs), in which the steady state circulation of the oceans may be deduced in response to a given external forcing, i.e., specification of sources and sinks at the ocean surface of momentum, energy, and water. The efforts and accomplishments are significant, but the ocean models lag behind atmospheric GCMs, primarily because the data for the oceans are far more sparse and the turnover constants span a much greater range. In addition, to employ ocean GCMs for studies of the carbon cycle, one needs to incorporate biological and chemical processes that occur within the fluid medium being modeled (i.e., the ocean).

From a long-term perspective, the development of ocean GCMs is undoubtedly the most rational approach not only for improving our understanding of the general circulation of the oceans but also for dealing with the issue of atmosphere-ocean carbon exchanges. However, these models have only recently begun to incorporate carbon dynamics; they demand significant computing resources; and they are rather complicated, employing a number of subroutines (and people). Thus, they may not yet be sufficiently well developed or sufficiently well understood to serve as the appropriate introductory

scientific tool for gaining insight into the ocean uptake of CO_2 . Simpler models are perhaps more appropriate to this pedagogical requirement; furthermore, simple models may be useful in the policy process.

An approach to studying the general circulation of the ocean, including carbon exchanges, that has been taken by chemical oceanographers offers a particularly useful framework for describing and understanding the basic outline for ocean-atmosphere carbon exchanges. For quite some time, they have been exploring several questions: What transfer processes due to water motions and biological activity are needed to explain the quasi-steady distribution of chemical compounds in the ocean? What can we learn about the biochemistry of the sea by the simultaneous use of several tracers? How can we extract information about the ocean from the observed transient tracers, particularly tritium, ^{14}C , and chlorofluorocarbons, that are now invading the world ocean due to human-caused emissions into the atmosphere?

The basic idea of this approach is to make use of the fact that the distributions of chemical and physical properties (e.g., concentrations of inorganic carbon or the varying pattern of temperature) are a reflection of the water motions (advective and turbulent) and perhaps of biological activities (e.g., new production, decomposition, carbonate formation, and dissolution) as well as the result of different distributions of sources and sinks.

For example, Craig (1957) recognized that the distribution of ^{14}C in the sea could be used to determine the rate of renewal of the deep waters of the oceans, and this led to the development of simple ocean models to interpret available ^{14}C data. The simultaneous use of ^{14}C and more traditional marine tracers was first attempted by Bolin and Stommel (1961) and further developed by Keeling and Bolin (1967). These early developments were based on very simple (usually two-box) ocean models. Oeschger et al. (1975) recognized the need to resolve the vertical structure of the oceans in more detail to depict reality better, and thereby formulated the box-diffusion model (Figure 2). Siegenthaler (1983) further elaborated on this same idea by considering the isopycnal ventilation of the deep sea through the cold polar regions (the outcrop model) (Figure 3). Björkström (1979) also incorporated cold polar regions which communicated directly with deep layers as distinct from warmer regions which did not. This simple low-dimensional box model of the global ocean (Figure 4) also distinguished intermediate waters from deeper ones and included advection as well as eddy diffusion. It also contained biological processes (organic and carbonate) as well as a full treatment of ocean surface carbon chemistry.

Box-Diffusion Model
(Oeschger et al., 1975)

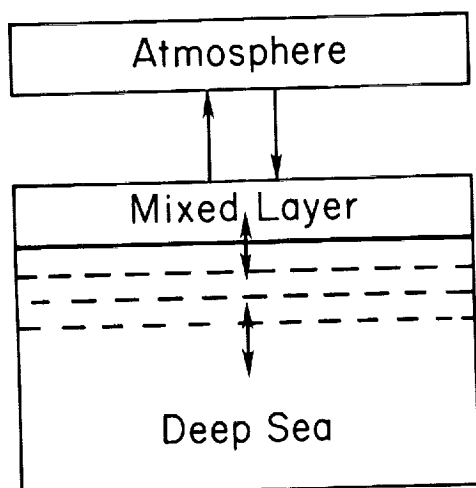


Figure 2. Box-diffusion model: the turnover of carbon below 75 m is represented by a diffusion equation. A constant coefficient of diffusivity is estimated to match an idealized profile of natural ^{14}C (Oeschger et al., 1975).

Box-Diffusion Model with Polar Outcrops
(Siegenthaler, 1983)

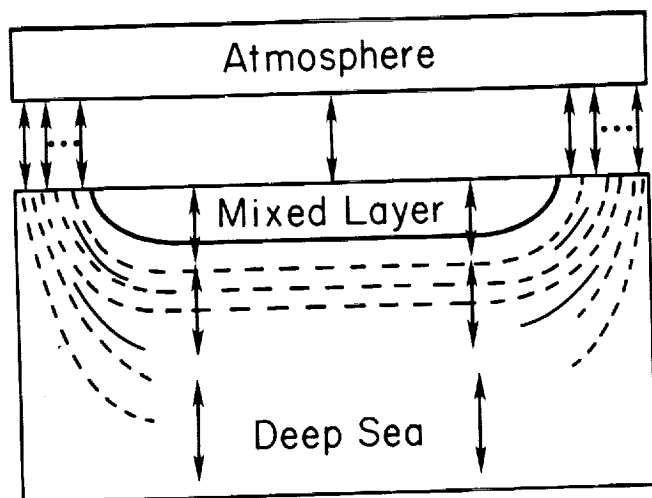


Figure 3. Outcrop-diffusion model: direct ventilation of the intermediate and deep oceans at high latitudes is allowed by incorporating outcrops for all sublayers into the box-diffusion formulation (Siegenthaler, 1983).

Advection-Diffusion Model
(Björkström, 1979)

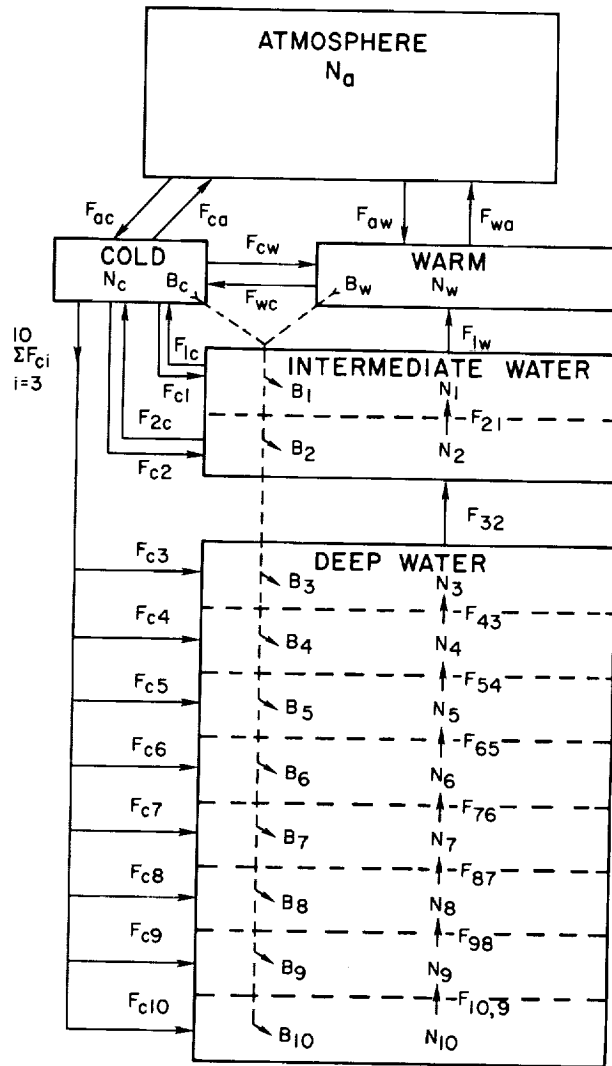


Figure 4. Advection-diffusion model: the first of a number of models whose structures are meant to be more realistic than the simple diffusive assumption. Boxes are labeled with an N with subscript, water mass fluxes are labeled by an F with subscript (some are in both directions and act as mixing terms); the biotic flux from the surface boxes to deeper layers (the biological pump) is denoted by B with the dashed vertical line, and the dissolution or decomposition within a nonsurface box is indicated by the solid arrow (Björkström, 1979).

Broecker and colleagues have made extensive use of tracers in the study of the biogeochemistry of the oceans, particularly by applying the concept of large-scale mixing of major water masses ("endmembers") with particular chemical signatures (e.g., Broecker, 1974; Broecker and Peng, 1982). When these formulations are expressed as continuity equations, the resulting matrical system is often mathematically indeterminate (i.e., there are an infinite number of solutions), and it is not generally clear how to select a single solution out of the many that are possible, particularly in light of the fact that one of the researcher's interests is the time-dependent aspects of the solution. Simply stated, it is difficult to know which of these indeterminate solutions to select or how to evaluate the response to a given forcing of the entire class of solutions.

In a deterministic setting, Bolin et al. (1983) have employed five tracers—total dissolved inorganic carbon, alkalinity, radiocarbon, oxygen, and dissolved inorganic phosphorus—in a constrained inverse methodology to parameterize a coarse-resolution (12 boxes) ocean model (Figure 5) in their study of the role of the ocean in the global carbon cycle. The fluxes of organic and inorganic detrital matter from the surface layers into the thermocline region and the deep sea were considered, and these fluxes became additional unknowns. The source and sink terms for the biologically important tracers were interrelated by assuming constant values for the Redfield ratios (the ratios of carbon to nitrogen to phosphorus to oxygen in organic matter). No deposition of detrital matter on the sea bottom was permitted. The inverse problem was formally overdetermined, and a condition of minimizing the errors of the conservation equations was employed, as well as a set of inequality constraints requiring that turbulent transfer be along the concentration gradient of the tracers and that detrital matter form in surface reservoirs and be dissolved or decomposed in deeper layers of the ocean. The deduced circulation pattern generally agreed with the current view of the gross features of the global ocean circulation. These same techniques were used to develop a more detailed model of the Atlantic Ocean (Bolin et al., 1987; Moore et al., 1989).

We should note that the choice of an indeterminate system, for example the Broecker and Peng models, or a determinant (often called incompatible) system, for example the 12-box model of Bolin and colleagues, is not a choice between bad and good models, but rather a more complicated issue of how much information can be extracted from a given set of observations. The issue goes beyond the scope of this paper and is addressed more fully in Fiadeiro and Veronis (1984), Wunsch (1985), Bolin et al. (1987), and Moore et al. (1989).

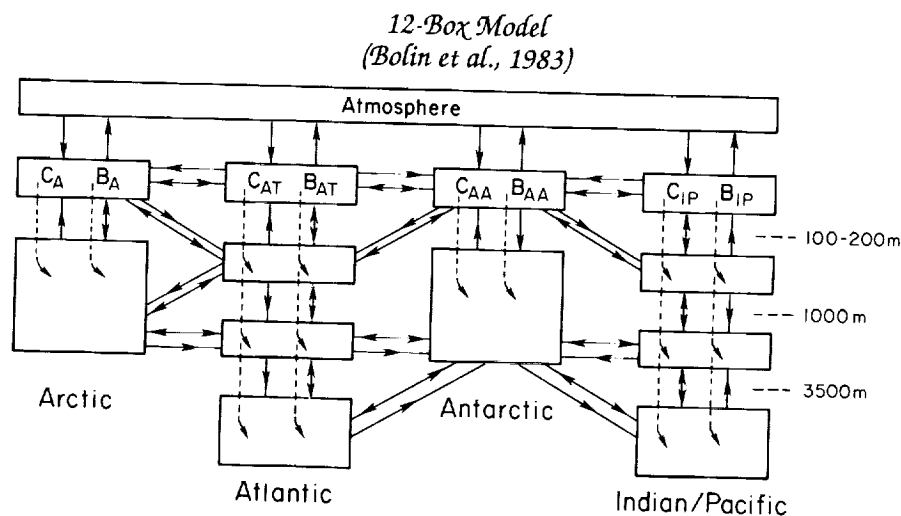


Figure 5. Twelve-box model: the Atlantic and Pacific-Indian oceans are each divided into surface, intermediate, deep, and bottom water components. The Arctic and Antarctic oceans are divided into surface and deep water components. The soft tissue and carbonate formation in surface boxes is indicated by B and C, with appropriate subscripts for region, and the fluxes to and decomposition in deeper layers are indicated by the vertical dashed line and the branching solid arrow, respectively. The model is calibrated against multiple tracer distributions (Bolin et al., 1983).

In considering the overall collection of simple, low-dimensional box models of the ocean carbon system, it appears that these models do provide insight into the processes, and their relative roles, controlling the ocean-atmosphere exchange of CO_2 .

Key Ocean Processes Controlling CO_2 Exchange

The oceans are a large sink ($2\text{--}3 \text{ Pg C/yr}$; $\text{Pg} = 10^{15} \text{ g}$) for the additional CO_2 that is being added to the atmosphere by human activities; however, the rate of carbon uptake by the oceans is limited by surface ocean chemistry and biology, and by the various patterns of mixing and circulation that determine the amount of carbon transported from surface waters to the deep ocean, where long-term carbon storage occurs.

The exchange of CO_2 between the ocean and the atmosphere is controlled primarily through gas transfer at the air-sea interface and hence is governed by the difference in partial pressure of CO_2 between the atmosphere and the sea surface and the rate of air-sea exchange. One way to view this exchange is to consider the two components separately: (1) the flux of CO_2 from the atmosphere to

the sea (F_{am}), and (2) the flux of CO_2 from the sea surface to the atmosphere (F_{ma}) (see Björkström; 1979).

The flux F_{am} can be represented as a first-order process dependent only upon the atmospheric residence time (T_a) of CO_2 (note that this is different from the atmospheric lifetime for CO_2 in the atmosphere, since a given molecule may move several times back and forth between the atmosphere and the ocean before being "sequestered" in the ocean or perhaps in the biosphere) and the mass of carbon as atmospheric CO_2 (N_a). This expression, and in particular the assumption about residence time, is an implicit statement about the globally averaged rate of air-sea exchange. The values of T_a and N_a are reasonably well established as global averages; however, the actual distribution of the air-sea exchange rates is quite complicated, depending on sea surface winds and sea surface state, and is not well established.

For the flux F_{ma} , let us consider a unit area (A) of ocean. Then the flux from the ocean to the atmosphere across A can be expressed as:

$$F_{ma} = (A/A_m)(N_{ao}/P_{ao})(P_A/T_a) \quad (1)$$

where A_m is the surface area of the oceans ($3.6 \times 10^{14} \text{ m}^2$), and N_{ao} is the mass of carbon in the atmosphere as CO_2 at some initial point in time with partial pressure P_{ao} ; the established value for N_{ao}/P_{ao} is $2.1314921 \times 10^{15} \text{ g C (ppmv)}^{-1}$. The key unknown is P_A , the partial pressure of CO_2 in the surface area A . Thus, the essential uncertainties are the distribution of the partial pressure of CO_2 in the surface of the ocean and the factors controlling this distribution.

From the perspective of the global carbon cycle one might consider the partial pressure of CO_2 in sea water as a linear function of the concentration of CO_2 in sea water (Broecker, 1974; Björkström, 1979); however, CO_2 dissociates in sea water at a rate that is itself a function of CO_2 concentration. Further complicating this chemical phenomenon are the dynamics of biological and physical processes. Primary production consumes CO_2 ; respiration and decay processes produce CO_2 . Each process affects the chemical equilibrium. Additionally, carbonate formation and dissolution alter alkalinity, which affects the partial pressure of CO_2 in sea water. The physical processes of advection and eddy diffusion continually alter not only the concentration of dissolved CO_2 in surface waters but the concentration of total inorganic carbon through the ocean, and thereby continually alter the CO_2 partial pressure distribution.

Although chemical, biological, and physical processes are combined in the marine environment, we shall discuss separately (and briefly) the principal attributes of each process.

Chemical Processes

Since CO_2 is not conservative in sea water, a more practical conceptual construct for determining the partial pressure of CO_2 in the ocean surface water is the concentration of total dissolved inorganic carbon (ΣC). This is defined as

$$\Sigma\text{C} = [\text{CO}_2] + [\text{HCO}_3^-] + [\text{CO}_3^{=}] \quad (2)$$

where "[]" indicates molar concentration. The difficulty with this construct is that the functional relationship between ΣC and the partial pressure P is only locally linear (depending not only on ΣC but on temperature salinity, and alkalinity as well), whereas P is essentially a globally linear function of $[\text{CO}_2]$. However, the conservative character of ΣC more than justifies the increase in complexity in calculating P and ΣC . In fact, one can treat the exploration as the answer to two questions: (1) What is the current relationship between ΣC and P ? and (2) How does this relationship change with future and perhaps far greater changes in ΣC ?

At a constant temperature, salinity, and alkalinity, and for relatively small changes in ΣC , the major aspects of the current relationship are captured by the linear function:

$$P = (10P_0/\Sigma\text{C}_0)\Sigma\text{C} - 9P_0 \quad (3)$$

where P_0 is the preindustrial partial pressure exerted by surface water and ΣC_0 is the sum of the concentrations of CO_2 , HCO_3^- , and $\text{CO}_3^{=}$ at an initial preindustrial point. Rewriting this linear expression as a normalized ratio of the change in P against the change in ΣC yields a more familiar expression:

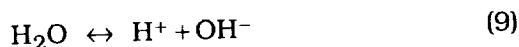
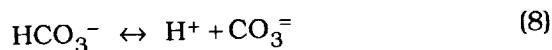
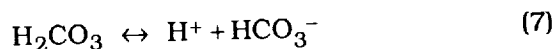
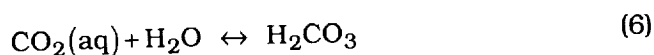
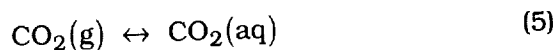
$$\{(P - P_0)/P_0\} / \{(\Sigma\text{C} - \Sigma\text{C}_0)/\Sigma\text{C}_0\} = 10 \quad (4)$$

This ratio, usually denoted R and termed the Revelle factor, was measured as part of the Geochemical Ocean Sections Study (GEOSECS) (see Takakashi, 1977), with a range from 8 for the warmest surface waters to 15 for the coldest. Considering the surface area of these waters, 10 is a good weighted average value. This expression, with a Revelle factor of 10, is used in the Oeschger et al. (1975) model (Figure 2) at the air-sea interface.

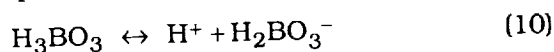
A direct calculation may clarify further the locally linear relationship expressed by Equation (4) and the essentially nonlinear behavior of P against ΣC , which is important in carbon cycling as CO_2 increases. Also, we need to consider formally the role of chemical solubility because it is temperature dependent. This calculation explicitly exploits ionic equilibria, and hence it describes more precisely the underlying relations between the partial pressure of CO_2 in the surface waters and its solubility at different temperatures,

ΣC , the concentration of hydrogen ions, and total alkalinity. A more thorough and yet easily grasped picture of carbonate-borate chemistry is given by Baes (1982).

The ionic composition of CO_2 in sea water is determined largely by the carbonate system and secondarily by the borate system. The underlying carbonate of sea water is expressed through the following equilibria reactions:



and the borate system is expressed by



If P denotes the partial pressure of CO_2 in sea water, then the ratio of the concentration of $\text{CO}_2(\text{aq})$ ($[\text{CO}_2]$) to P is the solubility, K_0 , of CO_2 . Similarly,

$$K_1 = [\text{H}^+][\text{HCO}_3^-] / [\text{CO}_2] \quad (11)$$

and

$$K_2 = [\text{H}^+][\text{CO}_3^{2-}] / [\text{HCO}_3^-] \quad (12)$$

are the first and second dissociation constants of carbonic acid (H_2CO_3), respectively. [Actually, K_1 is the apparent first dissociation constant for H_2CO_3 . In other words, it is the product of the inverse of the equilibrium constant for the hydration process of $\text{CO}_2(\text{aq})$ and the first dissociation coefficient of H_2CO_3 .]

A direct calculation yields the relationship between ΣC and $[\text{CO}_2]$:

$$[\text{CO}_2] = [\text{H}^+] \Sigma C / \{ [\text{H}^+]^2 + [\text{H}^+] K_1 + K_1 K_2 \} \quad (13)$$

Consequently, if one knows ΣC and $[\text{H}^+]$ then it is possible to calculate $[\text{CO}_2]$, P , and hence F_{ma} . However, with changes in $[\text{CO}_2]$ and hence ΣC , $[\text{H}^+]$ is altered through the carbonate reactions; furthermore, the production of carbonic acid and its dissociation perturbs the boric acid speciation in seawater, and hence $[\text{H}^+]$, since the carbonate ions compete with the borate ions for the available hydrogen ions. If one assumes that the carbonate-borate alkalinity, denoted by A , remains constant (Keeling, 1973), where A is defined by

$$A = [\text{HCO}_3^-] + 2[\text{CO}_3^{2-}] + [\text{H}_2\text{BO}_3^-] + [\text{OH}] - [\text{H}^+] \quad (14)$$

it is then possible to use the dissociation coefficients of H_2CO_3 , boric acid (H_3BO_3), and water to determine $[\text{H}^+]$, given only ΣC . In fact, given any two of the following four variables— P , ΣC , $[\text{H}^+]$, and A —the remaining two terms are determined.

In closing this section, let us return to Equation (4), the simple linear expression for the relationship between total dissolved inorganic carbon and the associated partial pressure—the Revelle factor. The importance of this factor is best expressed by an example that uses current data. Since the atmospheric concentration of CO_2 has increased by approximately 25% over the last 125 years since 1860, then in light of Equation (4), the increase in surface concentration of total dissolved inorganic carbon would be only 2.5% (assuming the surface layer remains in equilibrium with the atmosphere). Since this surface mixed layer is about the same size, with respect to total inorganic carbon, as the atmosphere (the 900 Pg C in Figure 1 reflects a somewhat deeper mixed layer than is generally accepted), this small increase would imply, for an abiotic, physically static ocean, much less carbon uptake than is generally associated with the role of the ocean in the global carbon cycle. Consequently, the biological and mixing processes must have removed from surface waters whatever additional CO_2 entered the ocean.

Biological Processes

Although the oceans are the largest active reservoir of carbon and cover almost 70% of the globe, the total marine biomass contains only about 3 Pg of carbon or just about 0.5% of the carbon stored in terrestrial vegetation. On the other hand, total primary production of marine organisms is 30–40 Pg/yr, corresponding to 30–40% of the total primary production of terrestrial vegetation. However, only a relatively small portion of this production results in a sink for atmospheric carbon, primarily through the sinking of particulate organic carbon, which decomposes in deeper layers or is incorporated into sediments. There is, in addition, a very large pool of dissolved organic carbon. The size of this pool is uncertain, but it appears to be on the order of the pool of carbon stored in terrestrial soils, approximately 1000–1500 Pg C.

The consumption of CO_2 in primary production in biologically active surface water, and the enrichment of the deep water in ΣC as the result of the decomposition and dissolution of detrital matter that originates from biological processes in the surface waters, taken together, are often referred to as the biological pump; the biology “pumps” carbon to the bottom. The most obvious pumping

is the incorporation into living organisms of ΣC that is dissolved in surface waters, either in tissue or as carbonate in shells. This lowers the partial pressure of CO_2 , followed by the "shipping" of some of the ΣC to the bottom, "packed" in the remains of dead marine organisms.

So, as a consequence of the biological pump, the concentration of total dissolved inorganic carbon is not uniform with depth; the concentration in surface waters is 10–15% less than deeper waters. There is a corresponding depletion of phosphorus (and nitrogen) in surface water, even in areas of intense upwelling, as the result of biological uptake and the loss of the detrital material, which also contains phosphorus (and nitrogen) as well as carbon.

The fate of the carbon that falls from the surface waters depends, in part, upon its characteristics. If it is organic material, then it is oxidized at intermediate depths, which results in an oxygen minimum and a carbon and phosphorus maximum. If the material is carbonate, it dissolves, raising both alkalinity and the concentration of carbon, primarily at great depths where the high pressure (and the effect of the increased "corrosiveness" of the sea water because of increased concentrations of ΣC) increases the solubility of calcium carbonate.

Thus, the biological pump lowers the partial pressure of CO_2 in surface waters and increases it in deep water not in contact with the atmosphere. It is as if the biological pump moves the partial pressure around in a way that allows CO_2 to work its way into the ocean. However, since the preindustrial carbon cycle was in quasi-steady state and since there is little reason to believe that the biological pump has changed over the last 300 years, its direct role in the perturbation problem that has been induced by human activities may be minimal. There is the possibility of an increase in the importance of the biological pump as the result of human-induced nutrient fluxes, but the low carbon-to-nutrient ratios in marine organic matter mitigate against this possibility; see Smith and Mackenzie, 1991, and Revelle, 1991. However, in a changing climate the role of the biological pump, including issues like ecosystem and alkalinity change, is open to question. The interested reader should also see Broecker, 1991; Smith and Mackenzie, 1991; Banes, 1991; Sarmiento, 1991; Revelle, 1991; Rintoul, this volume; and Longhurst, 1991, for a lively discussion of the biological pump and other carbon-related issues.

Physical Processes

The role of the oceans in the carbon cycle is much dependent on their rate of overturning (the meridional circulation) and mixing. In

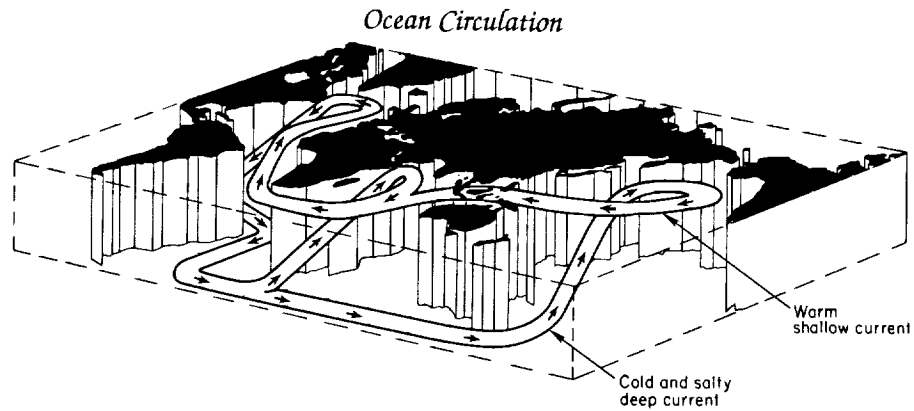


Figure 6. The broad meridional circulation of the ocean (adapted from Broecker and Peng, 1982).

polar regions, ice formation leaves much of the salt "behind," still in solution. The result is an increase in salinity in these already cold waters and hence an increase in density; furthermore, in the North Atlantic, evaporation exceeds precipitation. As a consequence, these cold, saline, and hence dense North Atlantic surface waters sink, and thereby they have the potential to form, in effect, a pipeline or conveyor belt (Figure 6; this concept was developed by Broecker; see Broecker and Peng, 1982) for transferring atmospheric CO_2 to the large reservoirs of abyssal waters which have long residence times. This downward convection of surface waters in polar regions of the North Atlantic during "bottom water formation" creates a sink for CO_2 in high latitudes, but the balancing upwelling of carbon-rich waters in low latitudes creates a source. In other words, what goes down (cold polar water with CO_2) must come up (warm equatorial water with excess CO_2). In addition to the bottom water formation in polar regions, there is water exchange between surface waters and intermediate waters due to vertical exchange in association with the surface ocean currents, like the Gulf Stream. These different exchange processes maintained by water motions renew the abyssal part of the oceans in a matter of a few hundred years in the Atlantic Ocean and up to about 1500 years in the Pacific. Intuitively we realize that this rather slow rate of oceanic turnover limits the oceans as a sink for CO_2 .

In order to sort out the effects of these many processes that govern the exchange of CO_2 between the atmosphere and the sea, it is necessary to use models. Such models range from complex formulations of the oceanic general circulation to simpler box models that seek to capture net effects of the processes.

Simple Ocean Carbon Models

Models are required to capture the net effect of these interwoven chemical, biological, and physical processes. As discussed above, GCMs that incorporate carbon dynamics are still under development and are rather demanding. In this paper, we shall explore briefly the four classical box models mentioned in the first section, which treat the oceanic uptake of CO_2 . We note in passing that one of these, the box-diffusion model, was used by the Intergovernmental Panel on Climate Change (IPCC) as the basis for determining future oceanic intake. The four models are:

- Box-diffusion model (BD; Figure 2): The turnover of carbon below 75 m is represented by a diffusion equation. A constant coefficient of diffusivity is estimated to match an idealized profile of natural ^{14}C .
- Advection-diffusion model (AD; Figure 4): The surface ocean is divided into cold and warm compartments. Water downwells directly from the cold surface compartment into intermediate and deep layers.
- 12-box model (12B; Figure 5): The Atlantic and Pacific-Indian Oceans are each divided into surface, intermediate, deep, and bottom water compartments. The Arctic and Antarctic Oceans are divided into surface and deep water compartments. The model is calibrated against multiple tracer distributions.
- Outcrop-diffusion model (OD; Figure 3): The intermediate and deep oceans at high latitudes are allowed direct ventilation by the incorporation of outcrops for all sublayers into the box-diffusion formulation.

The four models have, naturally, much in common: They are all diagnostic rather than prognostic, each uses ^{14}C in the parameterization process (in fact, ^{14}C is the basic clock for all of the models and hence controls much of their response), and each includes ocean carbon chemistry (buffer or Revelle factor) and some form of ocean mixing. There are, however, major differences: Ocean biology is explicitly included only in two (AD, 12B), whereas it is simply "part of" the diffusive process in both the BD and OD models; deep water formation is not explicitly considered in the box-diffusion model; the treatment of ocean chemistry varies in complexity; in the BD and OD models all of the physics is captured by a single constant eddy diffusivity term, whereas in the AD model the system is primarily advective (eddy diffusivities can be incorporated); the 12B model has both advection and eddy diffusivities; and, perhaps most impor-

tantly, the geometrical configurations are quite different. In addition, there are a host of smaller differences: ocean volume and surface area, parameterization procedures, ^{14}C profiles, initial conditions, and atmospheric residence time.

As a result of these large and small differences, we find (as described in the next section) a range of responses. Briefly, the OD model is the most efficient in taking up CO_2 ; the BD, AD, and 12B models all tend toward less carbon uptake. The reason the OD model is more efficient is because of the infinitely rapid connection of the atmosphere to the deeper layers; the reason the three other models are so similar is the overarching importance of ^{14}C in setting the basic rates within the models, but without a direct link to deeper layers.

We should note (again, see the next section) that their relative responses are in part a function of the forcing term, and in this study we consider only past and various future fossil fuel emission scenarios without any terrestrial release. Rapidly increasing emissions generally lead to a decreasing percentage going into the ocean model, whereas decreasing emissions have the opposite effect. Because they capture these changes, models are preferable to a constant multiplier or percentage for varying emission scenarios.

Basic Results

The change in atmospheric carbon dioxide is due primarily to two forcings: the combustion of fossil fuels and the net CO_2 flux to the atmosphere generated by changing land use patterns (Figure 7). The former is certainly greater today than in the past. There are significant uncertainties associated with the latter. For more discussion on calculating directly the biotic source, see Moore et al., 1981; Houghton et al., 1983, 1985, 1987, 1990, 1991; Palm et al., 1986; Detwiler and Hall, 1988; Melillo et al., 1988; Houghton and Skole, 1990; and Skole, 1990. We should also mention that perhaps the most careful analyses of the current characteristics of the carbon cycle that can be deduced from atmospheric data and models are provided by Keeling et al., 1989a, 1989b; Heinmann and Keeling, 1989; and Heinmann et al., 1989.

For the purposes of this paper, a more simplistic and direct consideration is appropriate. Consequently, we shall not consider the biotic flux in Figure 7, but simply the behavior of the four atmosphere-ocean systems when forced by the historic fossil fuel flux alone, except in one test case. In order to further simplify the comparison, we set all initial conditions at the 1860 ice core value for atmospheric CO_2 . The four ocean-atmosphere systems show

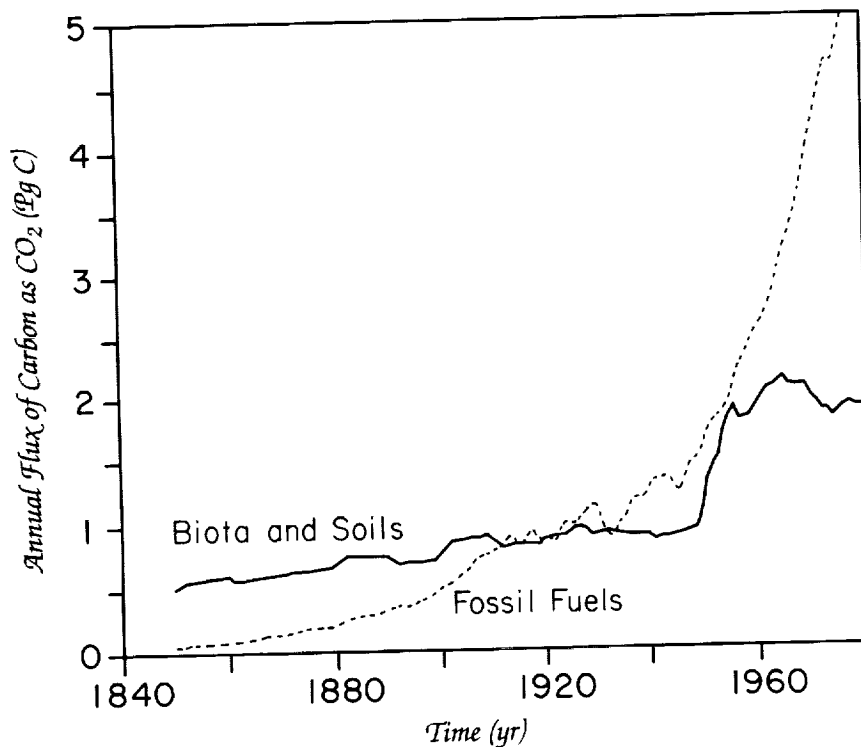


Figure 7. The fossil fuel emissions record (dashed line) and an estimate of the biotic release due to land use change (solid line). The fossil fuel record is from R. Rotty and G. Marland, cited in Moore et al. (1981). The biotic release is from Houghton et al., 1983.

remarkable agreement when compared to the ice core record for atmospheric CO₂ (Figure 8). One notes that if we had initialized the models at the 1740 ice core value, it would be possible to add additional CO₂ to the atmosphere during the early part of the record; but little, if any, additional CO₂ could be added to the current fossil fuel forcing term without exceeding atmospheric measurements (or introducing a new sink). We explore this in a highly simplified fashion simply to illustrate this point by defining an additional nonfossil fuel release (a "biotic" release; Figure 9), and then examine the response of the four ocean-atmosphere systems to the combined forcing of the historic fossil fuel record plus the "created" additional release (Figure 10). Obviously, the combined flux of fossil fuel with the calculated CO₂ flux associated with land use change (Figure 7) would take the response well above the observed increase in the atmosphere.

We now consider three future records simply to capture the general characteristics of the response of the ocean-atmosphere to dif-

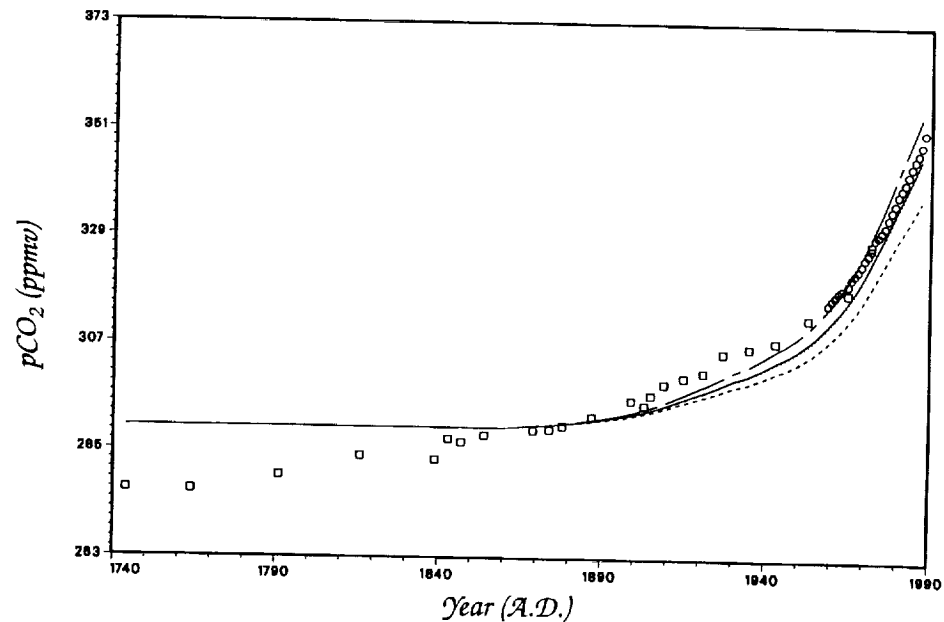


Figure 8. The responses of the four ocean-atmosphere models to the historic fossil fuel forcing. All models were initialized at the 1860 ice core value for CO_2 . Open boxes represent the CO_2 values from the ice core record at Siple Station, Antarctica, and the open circles at right are annual averaged atmospheric measurements from Mauna Loa Observatory (the Keeling record). The short dashed line represents the OD model results, the dotted line (just below the solid line) the AD model, the solid line the 12B model, and the short-long dashed line the BD model.

ferent CO_2 forcing; later, we will consider another slightly more policy-oriented forcing term. These initial three emission scenarios do not reflect any particular energy future; rather, they are chosen to reflect a broad range of responses. The two forcing functions are:

- The actual fossil fuel record (i.e., Rotty record; see Figure 7) to 1986 and then the best fit straight line to the 1977–86 data for the period 1987–2000 (Figure 11).
- A Gaussian-like fit (see Bacastow and Björkström, 1981) to the Rotty record such that (1) it has the same total input for the period 1860–1986, (2) it passes through the 1986 data point, and (3) the total integral is eight times the amount of CO_2 as in the preindustrial atmosphere (Figure 12).

We use the four models discussed earlier and actually include two parameterizations for the OD model. One uses steady-state ^{14}C to parameterize the dynamic processes in the model, and the other uses ^{14}C produced by nuclear bomb testing (see Siegenthaler, 1983). Hence, for each forcing term, there are actually five

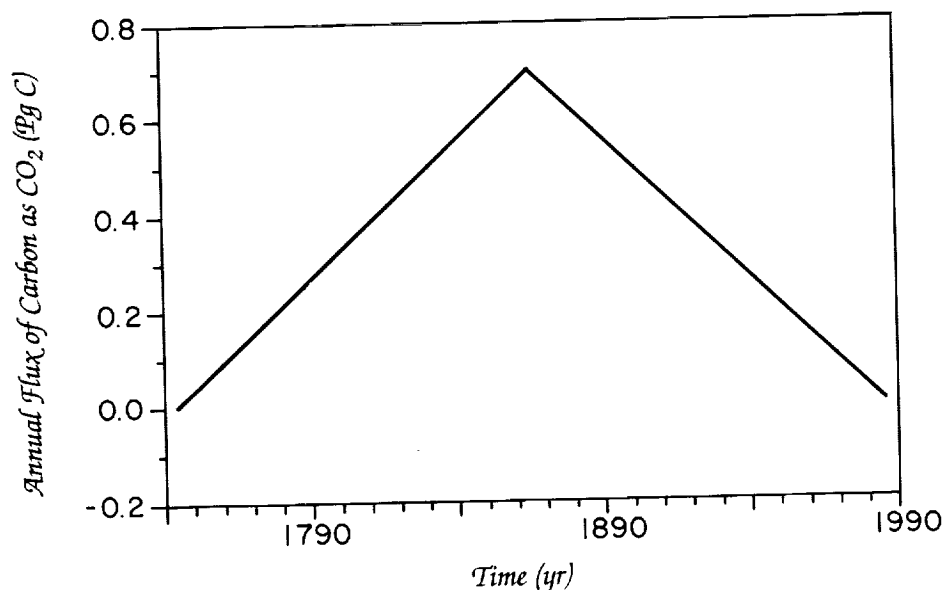


Figure 9. A highly artificial net biotic source flux of CO_2 .

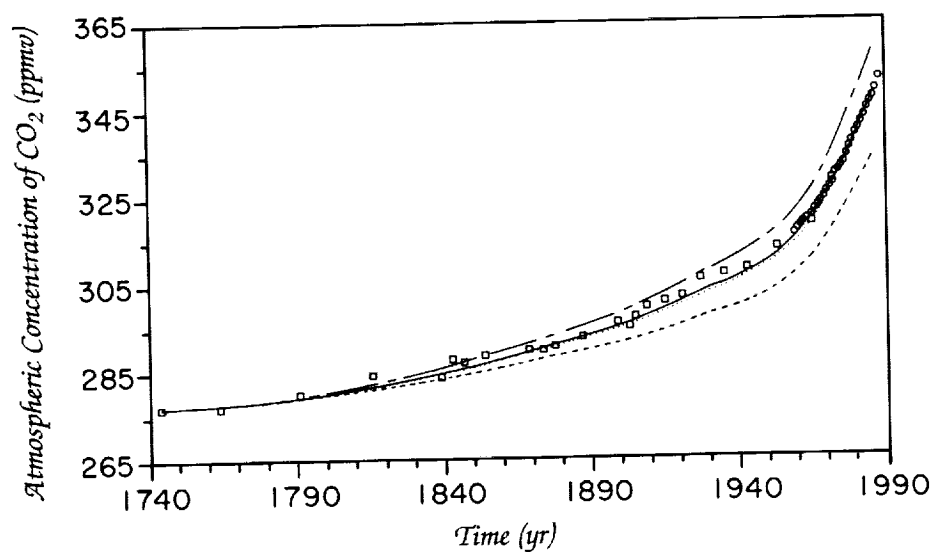


Figure 10. The response of the four ocean-atmosphere models to the historic fossil fuel forcing plus the source term shown in Figure 9. The models were initialized at the 1740 ice core value for CO_2 . Open boxes and open circles represent the Siple Station and Mauna Loa records, as in Figure 8. The short dashed line represents the OD model results, the dotted line (just below the solid line) the AD model, the solid line the 12B model, and the short-long dashed line the BD model.

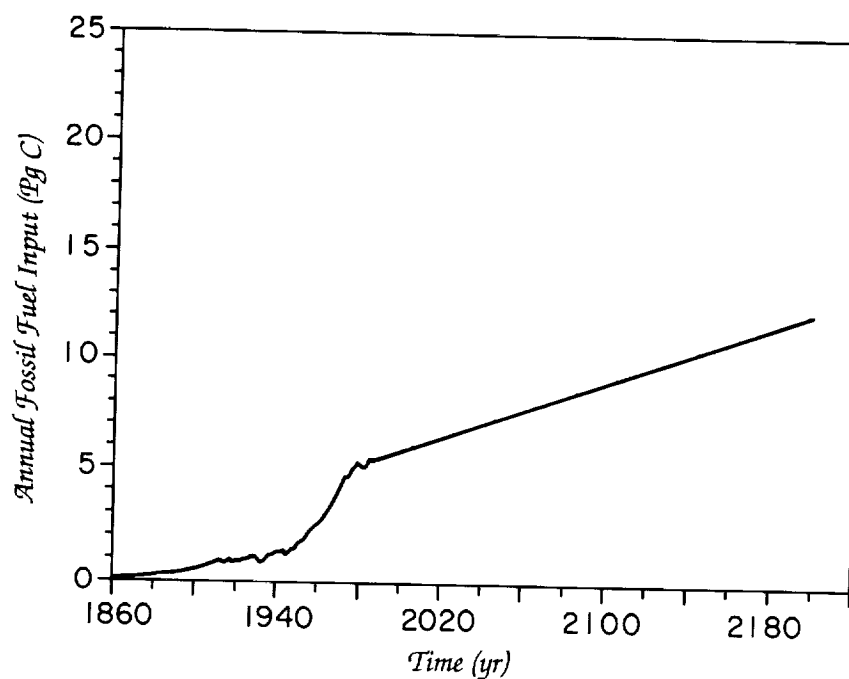


Figure 11. The fossil fuel record to 1986 and then the best fit straight line to the 1977-86 data for the period 1987-2000.

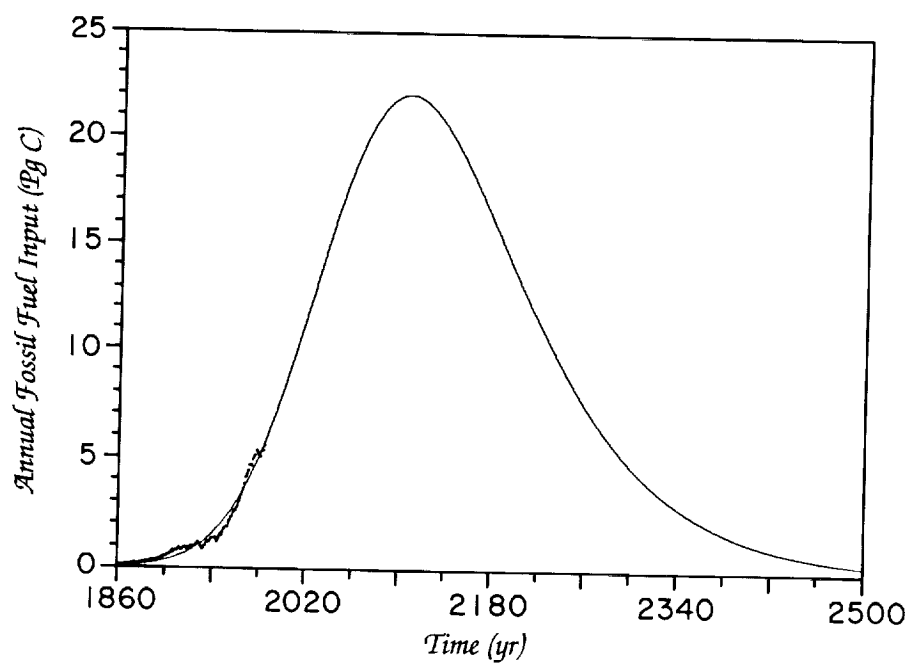


Figure 12. A Gaussian-like fit to the fossil fuel record.

responses. The results are shown in Figures 13–14. In using the Gaussian-like input functions, we extend the time frame to 2500 in order to show how the atmosphere-ocean system begins to relax as the forcing term (Figure 12) goes to zero. In all cases, the OD model, in either parameterization, is by far the most efficient in taking up CO_2 . The primary reason is that the structure of the OD model allows, as mentioned, an infinitely fast (and perhaps unreasonable) transfer of some of the atmospheric CO_2 to deep layers.

This range of models certainly includes what is the currently accepted role of the ocean in the global carbon cycle; however, it does not include a model that is efficient enough to allow for the uptake of the currently accepted deforestation-produced CO_2 , though the OD model comes the closest. Therefore, either (1) the current understanding of the oceans is insufficient, (2) the current estimates of land use-derived CO_2 are in error, or (3) there are important missing processes. It is likely that all three are true. One intermediate approach in testing the sensitivity of CO_2 projections is to include the mismatch as a “donor-controlled” sink in atmospheric concentrations when one includes as a forcing term the deforestation-produced CO_2 with the fossil fuel-derived CO_2 .

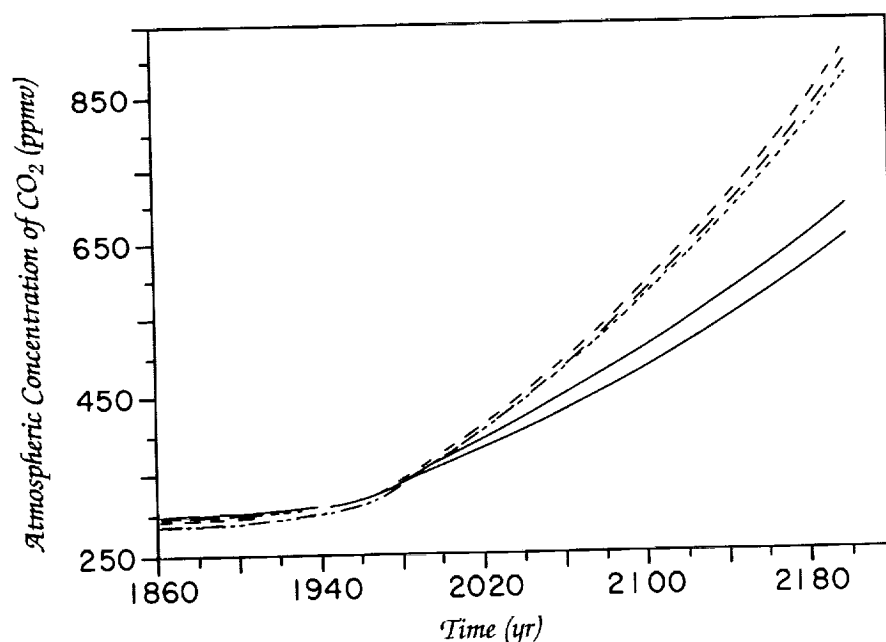


Figure 13. The responses of the ocean-atmosphere models to the fossil fuel forcing followed by the linear forcing (Figure 11). The solid lines represent two versions of the OD model, the short dashed line the AD model, the medium dashed line the BD model, and the long dashed line the 12B model.

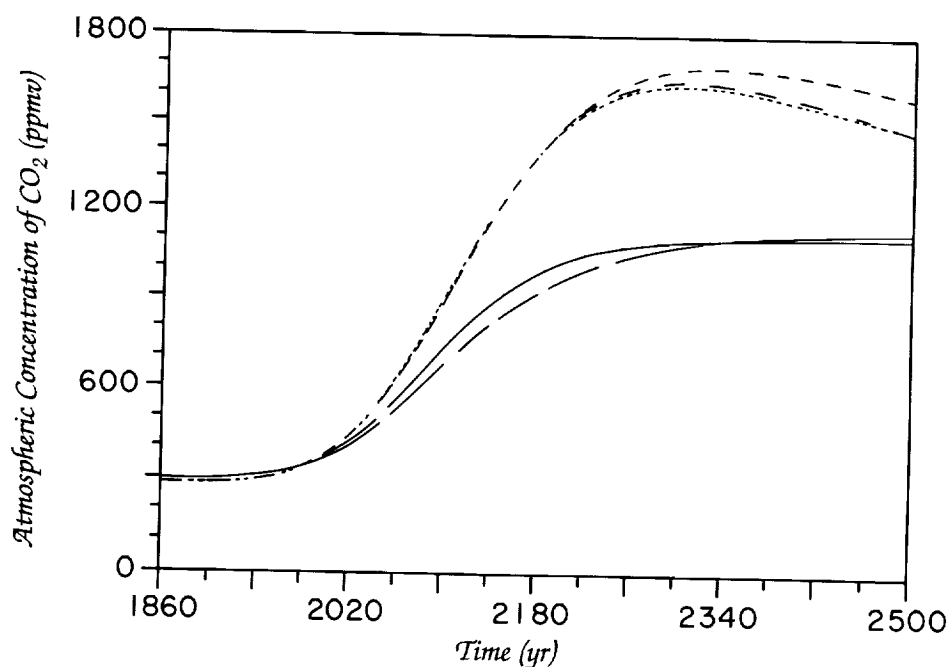


Figure 14. The responses of the ocean-atmosphere models to the Gaussian forcing (see Figure 12). The solid and dashed-solid lines represent two versions of the OD model, the short dashed line the AD model, the medium dashed line the BD model, and the long dashed line the 12B model.

Finally, in light of a recent editorial by Firor (1988), we tested the four models using an emission scenario investigated by Maier-Reimer and Hasselman (1987), the results of which are referenced by Firor. In this scenario, fossil fuel use decreases at 2% per year until it reaches half the current (1986) level of 5.56 BMT/yr; a graph of this forcing function is given in Figure 15. In his editorial, Firor suggests that society may be able to "come close to stabilizing the atmospheric burden of CO_2 with a 50% reduction in fossil fuel use."

The responses of the four models to this emissions scenario are given in Figure 16, and although there is an initial stabilization of atmospheric CO_2 , we find that after the input function becomes constant, the rise in atmospheric CO_2 concentration averages between 0.4 ppm/yr (for the OD model) and 0.5 ppm/yr (for the BD model). Admittedly, this rise is considerably less than the approximately 1 ppm/yr rise that is presently occurring, but it does not constitute a stable atmosphere. Tests of the responses of the four models to similar emissions scenarios indicate that emissions would have to be cut to approximately one-tenth of the present level to bring about something close to stabilization (a 0.05 ppm/yr rise).

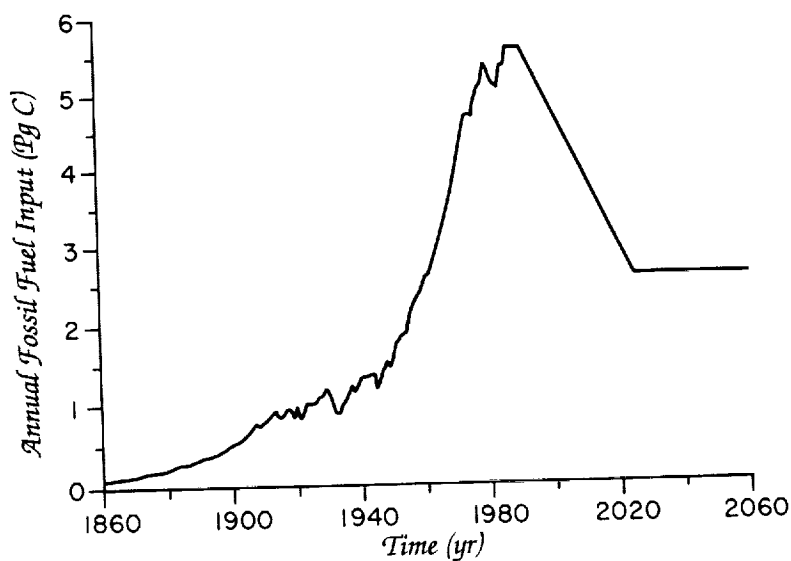


Figure 15. A plot of fossil fuel CO_2 emission to 1986 followed by a five-year period at the 1986 level, and then decreasing at 2% per year until it reaches half of the 1986 level of 5.56 Pg C/yr.

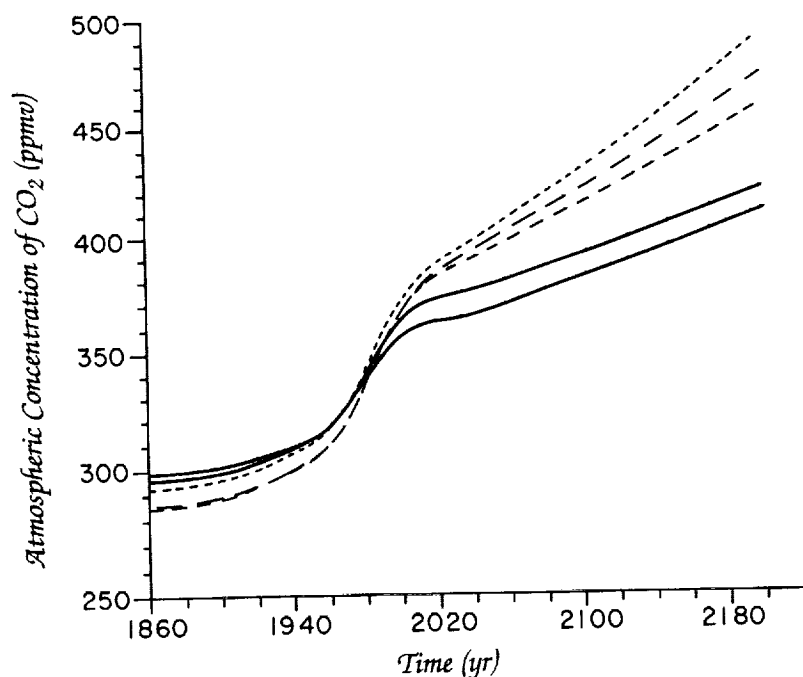


Figure 16. The responses of the four atmosphere-ocean models to the historic fossil fuel forcing, followed by a declining forcing as shown in Figure 15. The solid lines represent two versions of the OD model, the short dashed line the AD model, the medium dashed line the BD model, and the long dashed line the 12B model.

This topic and other related questions raised by Firor (1988) have been the subject of additional investigations (Harvey, 1989), and the differences between these studies in which atmospheric stabilization is achieved and the current findings need further investigation. This is but one area where simple models of the ocean-atmosphere carbon system might continue to prove useful. Other interesting issues relate to defining better the character of the terrestrial sink that would be required given an ocean-atmosphere model, a set of anthropogenic release terms, and the historical record for either CO_2 or ^{13}C .

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